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Detecting the long-term impacts from climate variability and increasing water consumption on runoff in the Krishna river basin (India)

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Abstract

Variations in climate, land-use and water consumption can have profound effects on river runoff. There is an increasing demand to study these factors at the regional to river basin-scale since these effects will particularly affect water resources management at this level. This paper presents a method that can help to differentiate between the effects of man-made hydrological developments and climate variability at the basin scale. We show and explain the relation between climate, water consumption and changes in runoff for the Krishna river basin in central India. Runoff under climate variability and increasing water consumption for irrigation and hydropower is simulated for the last 100 years using the STREAM water balance model. Runoff under climate variability is shown to vary only by about 14–34 mm (6–15%). It appears that reservoir construction after 1960 and increasing water consumption has caused a persistent decrease in annual runoff of up to approximately 123 mm (61%). Variation in runoff under natural climate variability only would have decreased over the period under study, but we estimate that increasing water consumption causes about two thirds of the current runoff variability.

1 Introduction

Human induced climate change, as well as natural climate variability, may have profound impacts on freshwater resources in many areas (Arnell et al., 2001). However, these impacts may be obscured by non-climatic factors, often anthropogenic in origin. Therefore, the relative impact of climate compared to non-climatic factors is important when studying the relation between climate and water resources availability. Non-climatic factors may be land use and land cover change and in particular developments in water storage in reservoirs and consumption for irrigation, drinking water and industry causes increased evaporation and substantial effects on runoff. For example, the global amount of water consumed for agriculture has roughly doubled between 1900 and 1980 (Falkenmark and Lannerstad, 2005). Water has therefore been identified as a critical factor for reaching the Millennium Development Goals (Rockström et al., 2005), and further assessment of shifts in water availability is needed.

Several studies have been devoted to either the impact of climate conditions or environmental and human use on water availability (e.g. Aerts and Droogers, 2004). Using hydrological models, it is possible to make a distinction between pristine catchment conditions and the effects of environmental changes (e.g. Letcher et al., 2001). Recent global studies on the effects of water storage and consumption have shown dramatic effects on the frequency of low flows and downstream water resources and services (Syvitski et al., 2005; Nilsson et al., 2005). Examples include the reduction of the amount of total runoff, the reduction in peak flow intensity, reduction in sediment transport, and changes in water quality, with consequences for downstream river morphology and ecology. Regional studies show similar trends. For instance, Magilligan et al. (2003) estimated that the peak discharges occurring every two years have decreased by about 60% for a number of river basins in the United States. Schreider et al. (2002) showed that due to the construction of small farm dams in Australia small but detectable changes can occur in the daily discharges. It has thus been argued that natural processes are no longer the sole influence on river systems: anthropogenic influences currently dominate (Meybeck, 2003).

Some researchers have approached these anthropogenic influences by using the green- and blue water concept. Green water refers to the amount of available freshwater that is used for evaporation in natural or agricultural vegetation, which is consumptive use, whereas blue water refers to the amount of water that is unaffected or remains as return flow. The blue water flow is important for downstream water availability, and it has been proposed that a certain requirement for minimum flow exists for ecological sustainability (Tharme, 2003). However, while an assessment of "green" and "blue" water flows is important for proper decisions in water resources management, the total amount of available freshwater from which allocations can be made is not constant over time, mostly because of variations in climate.

It appears, however, that very few studies pay attention to the combined effect of natural climate variability, climate change and anthropogenic impacts (e.g. Changnon and Demissie, 1996). It also happens that studies into water availability have used relatively short time intervals or concentrate on the average climate state and effects at the global or regional scale (e.g. Alcamo et al., 1997). Vörösmarty et al. (2000) compared the impacts from climate change and population growth and concluded that average climate change is likely to have a minor impact on water resources. However, they ignored the potential impacts that changes in year-to-year variability of climate may have. Most trend detection analyses so far have focussed on the analysis of the mean runoff and not on changes in runoff variability (for an overview see Kundzewicz and Robson, 2004). The assessment of historic high and low flows as demonstrated by Burn and Hag Elnur (2002), or statistical analyses applied to climate change scenarios as demonstrated for low flows by Arnell (2003), have shown the impact of climate variability on the variability of runoff. Studies of runoff effects caused by both climate variability and basin developments should consider long and discrete periods, preferably more than 50 years in order to capture multi-decadal variability of climate and runoff.

The main goal of the present research was to develop and test a method to separate the relative impact of variation in climate versus anthropogenic changes on runoff at the river basin scale. We have limited ourselves to studying the impacts of increasing water consumption for irrigation and hydropower on the annual and seasonal runoff over a period of 100 years in the arid region of the Krishna river basin located in central India. The objectives of this study were to:

- Assess and present statistics of the variation in climate and river discharges, in particular changes in precipitation and annual runoff;
- Calibrate and validate a spatial hydrological model in order to simulate monthly runoff over a 100-year period under climate variability, with and without accounting for changes in water consumption;

– Quantify changes in annual and seasonal runoff and runoff variability over 100 years by comparing observed and modelled monthly runoff;

– Determine the relative influence of variation in climate versus increasing water consumption on annual basin runoff and runoff variability.

2 Study area and data

2.1 The Krishna river basin

The Krishna river basin is the second largest river in peninsular India and stretches over an area of 258 948 km^2. The basin is located in the states of Karnataka (113 271 km^2), Andhra Pradesh (76 252 km^2) and Maharashtra (69 425 km^2). The basin represents almost 8% of surface area of the country of India and is currently inhabited by 67 million people. The major tributaries of the river include the Bhima River in the north and the Tungabhadra River in the south (Fig. 1). The river terminates at the Krishna delta in the Bay of Bengal. The climate in the basin is characterised by sub-tropical conditions with considerable rainfall in the mountains of the Western Ghatts and arid conditions in the basin interior. Total annual rainfall today averages 835 mm, while the annual average temperature reaches 26.7°C. Rainfall over India is highly variable due to the intra-seasonal and inter-annual variability of the South-West monsoon (June to September) and the North-East monsoon (October to November), leading to alternating drier and wetter conditions on the Indian continent (Krishnamurthy and Shukla, 2000; Munot and Kothawale, 2000).

Failing monsoons have often resulted in considerable declines in water availability and consequently led to increasing political tensions between the states. One of the driest recent episodes in Central India occurred in 1972 (see Fig. 2). Over 100 million people in India were affected as crops failed (http://www.em-dat.net). In 1973 the water allocation between the three riparian states of Maharashtra, Karnataka and Andhra

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Pradesh was settled in a water disputes act. Declines in water availability also impact on water quality. Chloride concentrations in the Krishna River, for instance, are highly correlated to total amounts of runoff (Sekhar and Indira, 2003). It has also been shown that sediment loads of the Krishna River have decreased over time (Ramesh and Subramania, 1988).

For many centuries small reservoirs, locally known as tanks, have been constructed to conserve and utilise water, and under British rule new canals were created, old tanks restored and new tanks built (Wallach, 1985). But the major reservoirs and canal systems now present in the basin were constructed during the second half of the 20th century for irrigation purposes and hydropower generation. Since the independence of India in 1947 the construction of reservoirs started to take off rapidly (Wallach, 1984). All large reservoirs with a storage capacity of more than 10^9 m^3 were built after 1953. The locations of the eight largest reservoirs in the basin are depicted in Fig. 1. These reservoirs were constructed between 1953 and 1988, and together they account for 26.6×10^9 m^3 or 80% of the capacity of large reservoirs in the basin. The storage capacity in the Krishna river basin is exceeded in India only by the capacity in the Ganges river basin. The benefits of water storage and redirection are clear: the current area of land that is being irrigated amounts to about 3.2×10^6 ha and a total of 1947 MW of electricity are produced annually.

2.2 Climate and runoff data

Climate data were retrieved from the global TS 2.0 dataset from the Climatic Research Unit, which covers the entire world for the period 1901–2000 on a 0.5 by 0.5 degree grid (Mitchell and Jones, 2005). Although this climate data has not been corrected for ambient factors, such as urban development or land use change, it is the most comprehensive climate dataset presently available and previous versions have often been used for studying the hydrological cycle.

Data on average monthly river discharges were taken from the RivDIS database available at http://www-eosdis.ornl.gov/rivdis/STATIONS.HTM (Vörösmarty et al.,

1998) for the downstream station at the city of Vijayawada (Global Runoff Data Centre station number 2854300) close to the mouth of the river; see Fig. 1. The data covers the period 1901–1979, with no data during the period 1961–1964 and for the year 1975. Additional discharge data for the period 1989–1999 were collected from yearbooks of the Indian Central Water Commission.

3 Trends in climate, peak runoff and reservoir development

The climate data, discharge data and data on reservoir construction were investigated in order to assess what determines the runoff of the Krishna river basin. We considered periods of 15 years in order to be able to determine changes between a number of coherent climatic periods.

In Fig. 2 the temperature and precipitation anomalies in the Krishna river basin are given as deviations from the 15-year period of 1901–1915. During this period the average annual total amount of precipitation was 765 mm, while the average annual temperature equalled 26.0°C. Variations between years and decades can clearly be observed. The data indicates that the average annual temperature increased by about 0.7°C, from 26.0°C over the period 1901–1915 to 26.7°C over the period 1986–2000. Average total annual precipitation increased slightly, by 9% between the same periods, from 765 to 835 mm.

Observed discharge data were converted from cubic metre per second into runoff in millimetres, using the basin size as reported by Vörösmarty et al. (1998), in order to be able to compare the runoff with precipitation amounts. The storage capacity of reservoirs larger than 10^6 m^3 has increased considerably after 1953, as can be seen from Fig. 3. The major reservoirs in the basin account for a storage capacity of 34.5×10^9 m^3. An additional volume is present in numerous smaller tanks and barrages spread out over the area. The height of the annual peak discharge has decreased from about 1969 onward; when the seven-year moving average of the peak discharge drops below the long-term minimum (Fig. 3). The decreased downstream runoff coincides with the rapid increase in reservoir storage capacity during the 1950s and 1960s.

4 Estimating changes in monthly runoff

From Figs. 2 and 3, the question arises how much water would have be available without reservoir development, and what variation in both monthly and seasonal runoff can be detected. For these purposes, a water balance model was developed to simulate monthly runoff under observed climate variability and changes in water consumption. Variations of monthly and seasonal runoff are important for the planning and management of agriculture, irrigation and hydropower production.

4.1 The STREAM model

The STREAM model is a spatial water balance model based on the formulation of the RHINEFLOW model (Van Deursen and Kwadijk, 1993) that calculates water availability and runoff on the basis of temperature and precipitation data and a number of land surface characteristics. The STREAM model for the Krishna River uses geographical information system (GIS) data at a spatial resolution of 3 by 3 km and at a monthly time-step. The water balance is calculated for each grid-cell using a direct runoff, soil water and groundwater component (see Appendix A). The STREAM model has been successfully applied in various forms to climate and hydrology studies in a number of river basins with similar size and characteristics as the Krishna river basin (Van Deursen and Kwadijk, 1994; Aerts et al., 1999, 2000; Middelkoop et al., 2001; Winsemius et al., 2006). These studies have confirmed that a monthly time step is sufficient for detecting decadal, inter-annual and seasonal changes in the hydrological cycle, such as those caused by water consumption and climatic change. The spatial resolution of 3 by 3 km is sufficient to analyse large-scale patterns, as the basin is approximately 260 000 km^2 in size and since the climate data is limited to a spatial resolution of 0.5 by 0.5 degrees.

4.2 Calibration and validation

First, the model was calibrated and validated. We assumed a baseline period between 1901 and 1915 for which the model was calibrated. The calibration of the model involved the adjustment of a reduction factor that tunes the reference evaporation (see Eq. A5), a coefficient that determines the separation between groundwater and runoff (Eq. A2), and a recession coefficient that determines the delay of the groundwater flow (Eq. A4). The calibration involved the match to observed total annual runoff, as well as seasonal patterns. The performance of the model was tested at every stage using the efficiency coefficient R^2 from Nash and Sutcliffe (1970). After the model was calibrated for the period 1901–1915, the following five 15-year periods for which observed data were available were used to validate the model.

The model was able to closely match the observed average annual runoff (see Table 1). The model results for the period 1901–2000 are shown in Fig. 4, together with the observed runoff. By comparing the observed runoff with the simulated runoff for the remaining 15-year periods the model performance can be assessed. The model efficiency coefficient after calibration of $R^2=0.73$ for the period 1901–1915 indicates that the model is capable of reasonably estimating mean monthly runoff for a total of 180 months, in particular when taking into account the high degree of human intervention in the hydrological cycle in this basin. The coefficient of determination (r^2) between observed and simulated monthly runoff is 0.77 for the period 1901–1915 (n=180) and 0.75 for the period 1901–1960 (n=720). The efficiency coefficient $R^2=0.68$ for the period 1916–1930 is slightly lower than the coefficient for the calibration period (Table 1), but the performance of the model for the two following periods (1931–1945 and 1946–1960) is sufficient to assume the model is accurately describing the runoff during these periods ($R^2=0.71$ and $R^2=0.73$ respectively). During the last two simulated periods (1965–1979 and 1989–1999) the fit of the model to the observed data is not good, as can be seen from the negative model efficiency coefficients in Table 1.

Monthly maximum runoff is approached reasonably only in absolute terms as can be seen from Fig. 4 (bottom); the coefficient of determination (r^2) between observed and simulated maximum runoff is only 0.25 for the period 1901–1960.

Furthermore, one of the advantages of the STREAM model is that it can generate spatial output of different parameters. Figure 5 shows the effective precipitation for the extreme dry year of 1972 and the moderate wet year of 1988. Effective precipitation was calculated as the total annual precipitation minus actual evaporation. During 1972 very little water was available and in particular the basin interior was extremely dry. The five driest and wettest years in terms of effective precipitation, as simulated by the model, are listed in Table 2. The average amount of effective precipitation in the period 1901–2000 was 278 mm. The amount that was available in 1918 was only 35%, while the amount that was available in 1903 was 197% of the average amount. Clearly, variation in precipitation can lead to considerable changes in the amount of water that is available for vegetation and humans.

4.3 Impacts on average annual runoff and maximum monthly runoff

There is a clear deviation of the simulated runoff with respect to the observed runoff after 1960, for both the total annual runoff and the maximum monthly runoff (Fig. 4). Although there were very little changes in total annual precipitation, there is a clear reduction in annual average runoff of approximately 84 mm (41%) and 123 mm (61%) and a reduction in the maximum monthly runoff of approximately 29 (37%) mm and 40 mm (53%) over the periods 1965–1979 and 1989–1999, respectively (Table 1). These values were calculated by subtracting the observed runoff from the simulated runoff. A t-test was applied in order to determine whether there is a significant change in observed runoff during the period 1965–1979, relative to the period 1901–1960. It turns out that the mean annual runoff has significantly changed already during this period (test value t=7.214, t_{crit}=3.460, p<0.0001).

The hydrological model was able to simulate the relative changes in runoff variability over the 15-year periods, although in absolute terms the model overestimated the variability (Table 1). The coefficient of variation, defined as the standard deviation divided by the mean runoff, was estimated to be approximately 1.5 times higher than the observed coefficient for the periods between 1901 and 1960. The variability in annual runoff follows very closely the changes in the variability of the total annual precipitation, until the period 1946–1960. After this period the observed variability in runoff increased, and by an amount that is higher than would be expected on the basis of the model results that were forced by the variability in precipitation only. Instead of a coefficient of variation of approximately 0.29 and 0.24 over the periods 1965–1979 and 1989–1999, as simulated by the hydrological model, the coefficient of variation in runoff increased to 0.37 and 0.53 over the periods 1965–1979 and 1989–1999, respectively. Taking into account the overestimation of the model, the coefficient of variation may have only been 0.20 and 0.16 over the periods 1965–1979 and 1989–1999, respectively. It appears therefore that two thirds of the current variability in runoff is caused by the decline in total runoff.

4.4 Impacts on seasonal runoff

Next, we simulated the difference (residual) between observed and modelled runoff for the different monsoon seasons. In Fig. 6 the normalised difference between the simulated and observed annual runoff over the period 1901–1979 is plotted against time for the monsoon season (June–November) and the post monsoon (December–May). This normalised difference d was calculated as

$$d = \frac{q_{\text{obs}} - q_{\text{sim}}}{q_{\text{sim}}} \tag{1}$$

where q_{obs} is the simulated amount of runoff and q_{sim} the observed amount of runoff in a particular year.

A steady decline in runoff during the monsoon season started around the beginning of the 1960s (Fig. 6, top). During the period 1965–1979 on average approximately half the runoff that is simulated was actually observed. This decreased further to approximately less than a third on average. An all-time low occurred during the year 1995, when only 10% of the estimated runoff was observed. The opposite pattern can be seen for the post monsoon season. Overall, observed runoff during the post monsoon season increased relative to what is estimated by the model, except for the period 1970–1974 when very little of the available water reached the outflow point. During the period 1965–1979 on average 1.5 times more runoff is observed than is expected on the basis of the model simulation (Fig. 6). This increased further during the period 1989–1999 to about three times the simulated runoff. In the year 1992 ten times the simulated runoff was observed during the post monsoon season.

The difference between the simulated and observed runoff will reflect environmental impacts other than climate variability, since the variability in precipitation is accounted for in both the observed and simulated runoff. The difference is probably mainly due to the obstruction of the river channel by dams and increasing water consumption. The timing of the change in the normalised difference supports this, as it coincides with the increase in reservoir capacity in the basin, as seen in Fig. 3. During the monsoon season the reservoirs are filled, resulting in a decline in runoff. As more water is released from the reservoirs for irrigation purposes, an increasing amount of return flows at the lower end of the river during the post monsoon season can be seen. This is particularly clear for the period 1951–1960 in Fig. 6. The reservoirs, their operation and the increasing water consumption are reflected in an overall reduced and more variable outflow at the lower end of the river basin.

5 Accounting for increasing water consumption

In previous sections we discussed the model results that incorporated only climate variability and compared these with the observed record. We now attempt to simulate the impact of increasing reservoir development and associated water consumption on the runoff. Changes in water consumption were assumed to be reflected in the difference between the simulated and observed runoff, as explained in the previous section.

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We calculated the ratio between simulated and observed runoff over the period 1965–1979 and used these as attenuation factors. Next, we derived a function of reservoir development, by comparing the reservoir capacity in a particular year with the average reservoir capacity between 1965 and 1979. We used the equations

$$\text{for} \quad R_y \leq 1, \; q'_{\text{sim},i} = q_{\text{sim},i} - q_{\text{sim},i}(1 - f_j)R_y \tag{2}$$

$$\text{for} \quad R_y > 1, \; q'_{\text{sim},i} = q_{\text{sim},i} f_j \big/ R_y D \tag{3}$$

where R_y is the reservoir capacity in year y, normalised to the period 1965–1979, $q'_{\text{sim},i}$ is the adjusted simulated runoff in month i in millimetres, $q_{\text{sim},i}$ is the original simulated runoff in month i in millimetres, $f_j = \bar{q}_{\text{obs},j} / \bar{q}_{\text{sim},j}$ with average observed and simulated runoff in month j (j is 1 to 12) for the period 1965–1979 in millimetres and D is a damping factor.

We used a damping factor since it is expected that a certain increase in reservoir capacity will not result in a proportionate reduction in runoff, as part of the irrigation water is rerouted to the river channel as return flow. This is evident from the fact that although the reservoir capacity continued to increase approximately threefold relative to the period 1965–1979 (Fig. 3), runoff did not decrease to a third of the previous period (see Table 1). The value of the damping factor was set at 0.84, as this provided the best Nash-Sutcliffe efficiency coefficient values for the periods 1965–1979 and 1989–1999.

Figure 7 depicts the results of the simulation, incorporating the effect of reservoirs. The fit of the simulated runoff to the observed data is better, for both the total annual runoff and the maximum monthly runoff. The model efficiency coefficients have improved relative to the model without reservoirs. For the period 1965–1979 the model fits well (R^2=0.69), while for the period 1989–1999 the model has improved considerably (R^2=0.41) (see Table 3). The model still could not approach the increase in variability of the total annual runoff that is observed during the periods 1965–1979 and 1989–1999. The observed variability could be a result of factors that are not included in the model, such as reservoir operation and timing of irrigation.

The estimated amount of water that is additionally evaporated for irrigation and hydropower purposes is plotted in Fig. 8. Note that these amounts are additional to amounts extracted by tanks and reservoirs constructed prior to 1901, which are incorporated in the "natural" vegetation evaporation estimate. Until 1953 a negligible amount of water was deviated from the main river. During the period 1965–1979 an average of 78 mm (38% of simulated runoff) was extracted. This estimated amount compares well with the estimated decline of 84 mm that was reported in Sect. 4.3. Additional water consumption increased to 139 mm (68% of simulated runoff) during the period 1989–1999. This amount is higher than the estimated decline of 123 mm reported in Sect. 4.3. This difference is probably caused by the rough estimation of water extraction, based on reservoir capacity increase only, using Eq. (3). Variations in reservoir operation are not taken into account. The estimate of Sect. 4.3 using the difference between observed and simulated runoff for the period 1989–1999 of 123 mm may be more accurate.

6 Discussion and conclusions

The construction of reservoir capacity in the Krishna river basin during the second half of the 20th century has been considerable. Our analysis has shown that observed downstream runoff in the Krishna river basin exhibited a strong decline after 1960. At the same time, peak discharges decreased substantially.

Using a hydrological model we were able to simulate the pristine situation, as well as spatial aspects of components of the hydrological cycle. The runoff as estimated by the water balance model deviates from the observed discharges, in particular during the period after 1960. This difference is attributed to increasing water consumption. An analysis of the residual between simulation and observation shows that a structural decline in the total average annual runoff of 123 mm (or 61% of simulated runoff) over the period 1989–1999 can be attributed to factors other than climate variability or climate change. During the monsoon season a decline of an average 121 mm (67%) was observed and during the post monsoon season an average relative increase in runoff of 8 mm (296%) was observed over the period 1989–1999.

The increasing water consumption was also simulated using the record of reservoir construction and the water balance model results. From these data it is estimated that increasing water consumption for irrigation and hydropower has contributed to approximately 134 mm extra evaporation in the last 10 years (1991–2000), which is about 21% of total annual evaporation and 68% of runoff in the basin as simulated by the hydrological model.

Changes in precipitation due to climate variability alone resulted in very little variation in runoff during the period 1901–1960. Observed climate variability accounts for changes in runoff of up to approximately +34 (15%) and −14 mm (6%) during the period 1901–1960. Variability in runoff (coefficient of variation of 0.20 to 0.25) changed little over the period 1901–1960 in response to variation in precipitation. In fact, variability in precipitation appears to have decreased over time (coefficient of variation of 0.18 in 1901–1915 to 0.08 in 1989–1999). Without increasing water consumption, runoff would have remained the same over the period of study, and variability would have decreased to about one third of the current observed runoff variability that is at 0.53.

The changes in runoff of the Krishna river basin and its variability over the last century are therefore likely to be due only to human interference and not to climate variability. However, severe events, such as the drought in 1972, are a direct consequence of shortfalls in precipitation. Changes in future climate may therefore have far-reaching effects in downstream areas, when more frequent dry periods compound with structural declines in runoff as a result of increasing consumption upstream.

The results of our research imply that when analysing the impact of climate variability, and also in analysing the impact of climate change, other environmental changes can be equally or more important. It is possible, however, to account for such changes, using the methods described above. The model can also be used to estimate the sensitivity to future climate change, using scenarios. The methods are fairly simple, but can clearly separate between different environmental changes, such as reservoir construction and water consumption. Moreover, other methods that assess temporal changes in water consumption and evaporation also rely on the availability of data. Remote sensing methods that estimate variation in evaporation for instance, need satellite data that are available only since the late 1970s. Water balance models offer a useful tool to estimate changes before that point in time. One condition for using this method, however, is that a discharge record and a record of climate parameters of sufficient length are available that can be compared to model output. Future studies may need to take into account other changes, such as land use change, and changes in evaporation that were only roughly estimated in this study.

Appendix A

STREAM model formulation

The STREAM model calculates the water balance for each cell in a grid at a monthly time step according to a number of parameters (Aerts et al., 1999). Total runoff T is calculated as

$$T = R + M + B \tag{A1}$$

where R is direct runoff, M is snow melt, and B is the base flow origination from groundwater, all in mm per month.

The direct runoff R is calculated from the soil water balance S using a separation coefficient s_c:

$$R = S \cdot s_c \tag{A2}$$

The remaining amount of water from the soil water balance is redirected to the groundwater (TG), using

$$TG = S - R \tag{A3}$$

The base flow is calculated from the amount of groundwater GW stored using a recession coefficient r_c:

$$B = GW/r_c \tag{A4}$$

The soil water balance and actual evaporation are calculated for each month using the equations from Thornthwaite and Mather (1957). Actual evaporation is estimated from adjusted reference evaporation, using a crop factor k_c and a reduction coefficient F_{red} that acts as calibration factor:

$$ET_0' = ET_0 \cdot k_c \cdot F_{\text{red}} \tag{A5}$$

Reference evaporation is calculated from temperature, using the formulas from Thornthwaite (1948). FAO factors were used for adjusting the reference evaporation to different land-cover types using crop factors (Doorenbos and Pruitt, 1975). Land-cover classes were taken from the Global Land Cover Characteristics database Version 1.2, produced by the International Geosphere Biosphere Programme (IGBP). This dataset is based on NOAA AVHRR satellite observations from April 1992 to March 1993, which were classified to land-cover characteristics by Belward et al. (1999). Parameters for the maximum soil water holding capacity were taken from a global dataset compiled by the United States Department of Agriculture (available from http://www.nrcs.usda.gov/) with a resolution of 2 arc minutes (about 3.5 by 3.5 km).

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Table 1. Observed average amount of annual precipitation and its coefficient of variation, observed and simulated total average annual runoff (in millimetres), their coefficients of variation (*CV*) and model efficiency coefficients (R^2) for the different periods. *n* designates the number of months that were used to calculate the *CV* and R^2 of the runoff.

Period		1901–1915	1916–1930	1931–1945	1946–1960	1965–1979	1989–1999
Precipitation	[mm]	765	737	786	865	798	847
	CV	0.18	0.20	0.15	0.12	0.14	0.08
Mean runoff	Observed [mm]	208	213	207	255	120	80
	Simulated [mm]	208	178	207	250	205	204
	Observed *CV*	0.25	0.29	0.20	0.23	0.37	0.53
	Simulated *CV*	0.41	0.45	0.29	0.30	0.29	0.24
	R^2	0.73	0.68	0.71	0.73	−0.14	−2.74
	n	180	180	180	180	168	144
Peak runoff	Observed [mm]	77	71	68	88	50	35
	Simulated [mm]	83	60	72	97	79	75

Table 2. Top five of simulated annual effective precipitation (defined as precipitation minus actual evaporation) averaged over the Krishna river basin.

Driest		Wettest	
Year	Effective precipitation [mm]	Year	Effective precipitation [mm]
1918	97	1903	546
1972	97	1956	525
1920	98	1975	489
1985	141	1916	472
1987	142	1964	453

Table 3. Observed average amount of annual precipitation and its coefficient of variation, observed and simulated total average annual runoff (in millimetres), their coefficients of variation (CV) and model efficiency coefficients (R^2) for the different periods. n designates the number of months that were used to calculate the CV and R^2 of the runoff. The simulation incorporates an increase in reservoir capacity.

Period		1901–1915	1916–1930	1931–1945	1946–1960	1965–1979	1989–1999
Precipitation	[mm]	765	737	786	865	798	847
	CV	0.18	0.20	0.15	0.12	0.14	0.08
Mean runoff	Observed [mm]	208	213	207	255	120	80
	Simulated [mm]	205	172	195	217	127	65
	Observed CV	0.25	0.29	0.20	0.23	0.37	0.53
	Simulated CV	0.41	0.45	0.29	0.27	0.32	0.24
	R^2	0.73	0.68	0.73	0.73	0.69	0.41
	n	180	180	180	180	168	144
Peak runoff	Observed [mm]	77	71	68	88	50	35
	Simulated [mm]	82	58	69	87	55	27

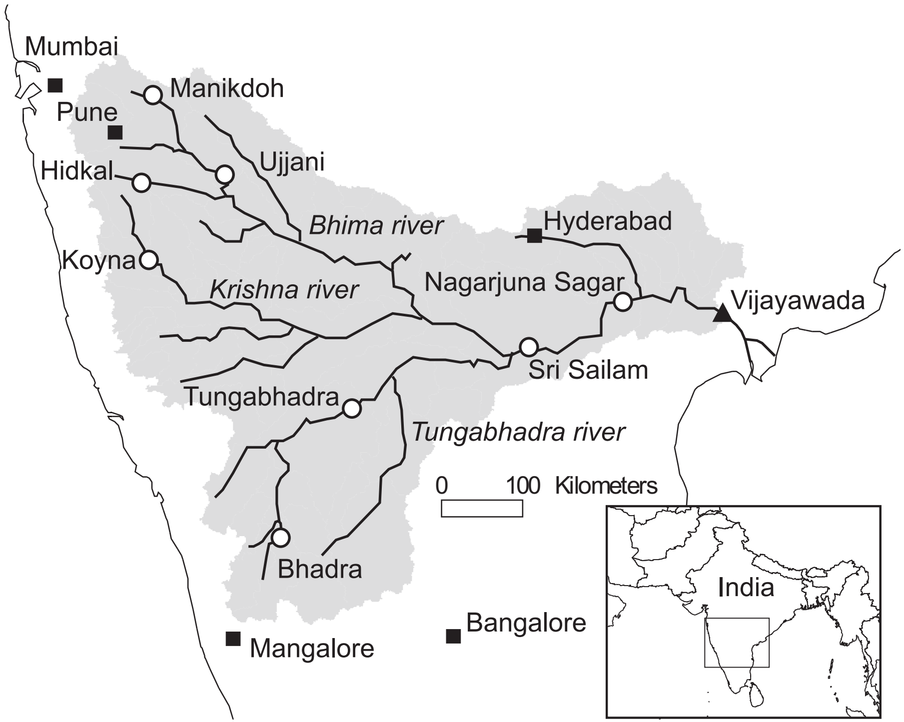

Fig. 1. Map of the Krishna river basin with its main tributaries, major cities and the discharge gauging station at Vijayawada (triangle) at the lower end of the river. The open circles indicate the locations of the eight largest reservoirs in the basin.

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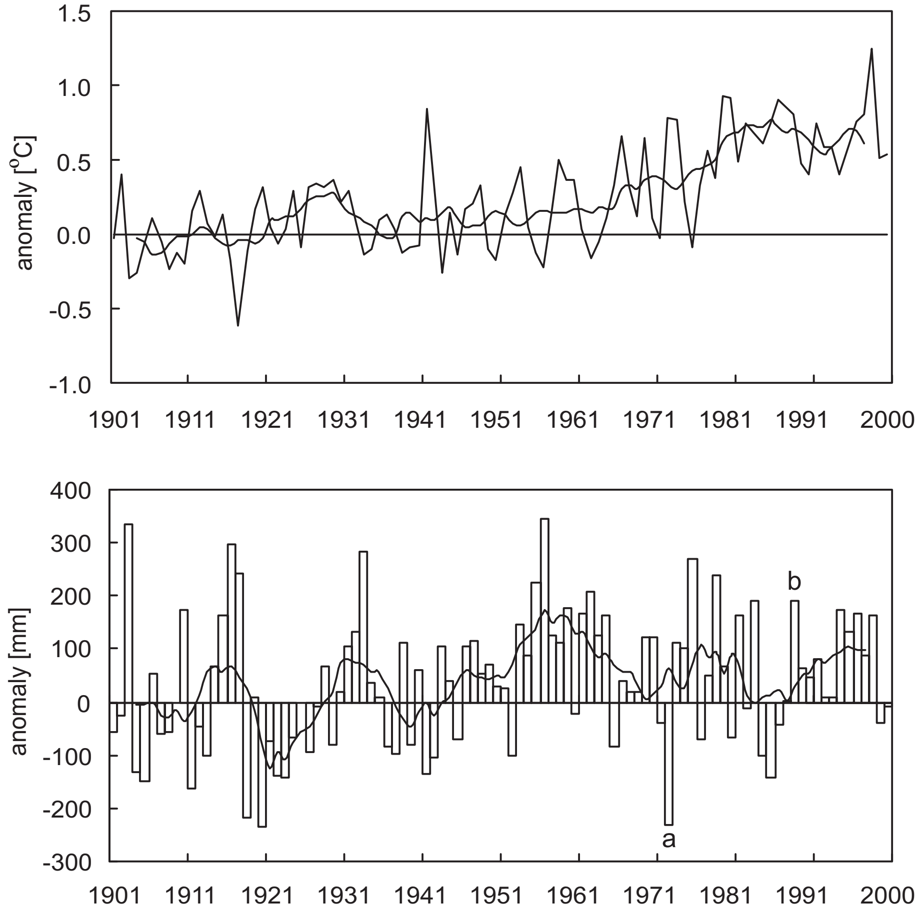

Fig. 2. Temperature (top) and precipitation (bottom) anomalies and their seven-year moving averages in the Krishna River Basin, relative to the period 1901–1915. In the lower graph, a and b designate a dry and a wet year, for which the spatial patterns of effective precipitation are plotted in Fig. 5.

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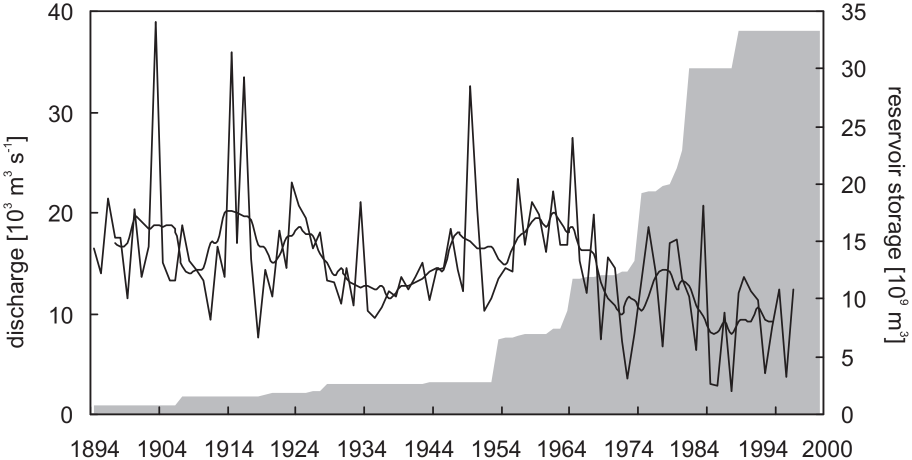

Fig. 3. Instantaneous annual downstream peak discharge values over the period 1894–1996 (line), its seven-year moving average and the cumulative reservoir storage capacity (shaded area) of reservoirs larger than 10^9 m^3 in the Krishna river basin over the period 1894–2000.

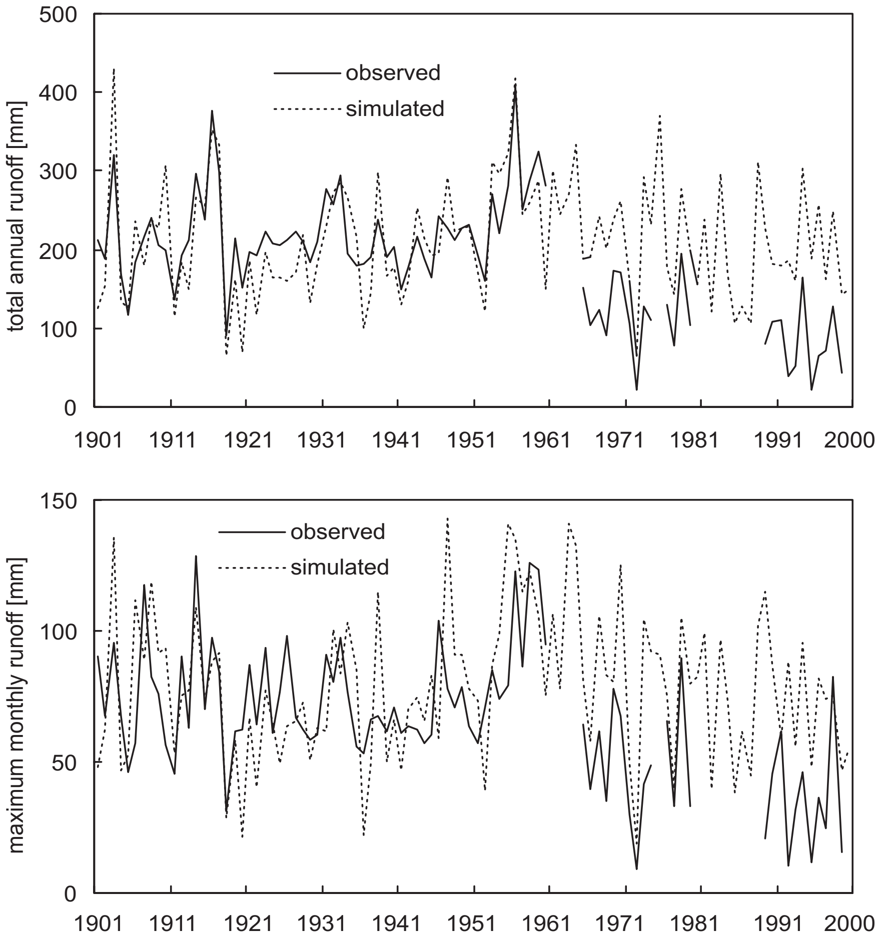

Fig. 4. Simulation model results for the period 1901–2000 compared to the observed total annual runoff (top) and maximum monthly runoff (bottom).

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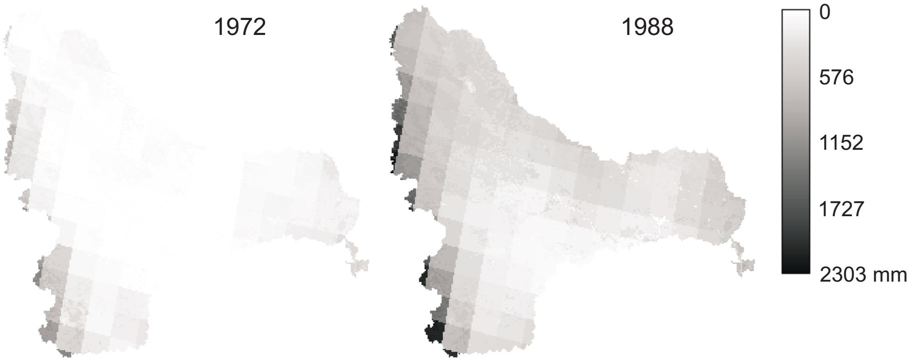

Fig. 5. Effective precipitation for the years 1972 and 1988.

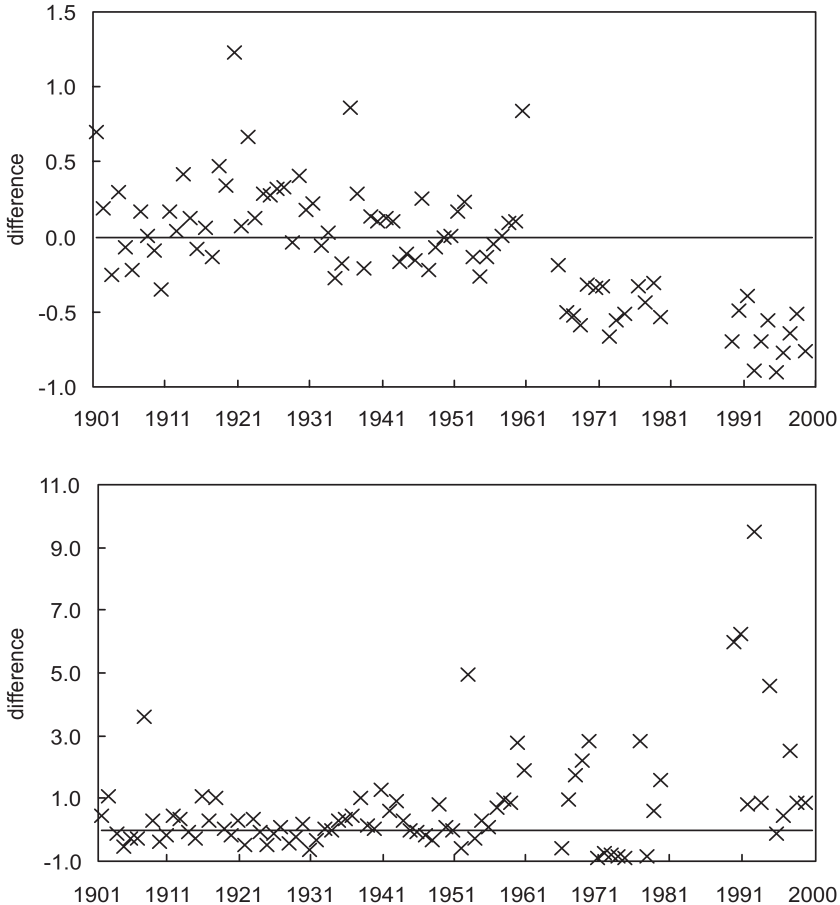

Fig. 6. Normalised difference between the observed and simulated runoff during the monsoon season (top) and the post monsoon season (bottom).

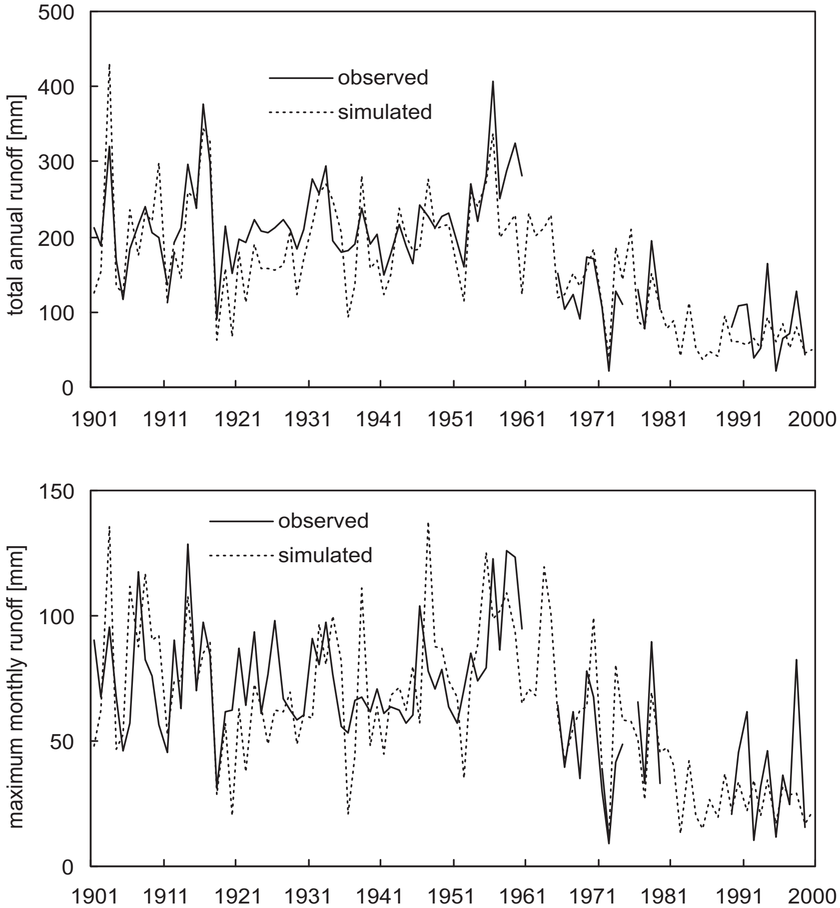

Fig. 7. Simulation model results, incorporating increasing water use, for the period 1901–2000 compared to the observed total annual runoff (top) and maximum monthly runoff (bottom).

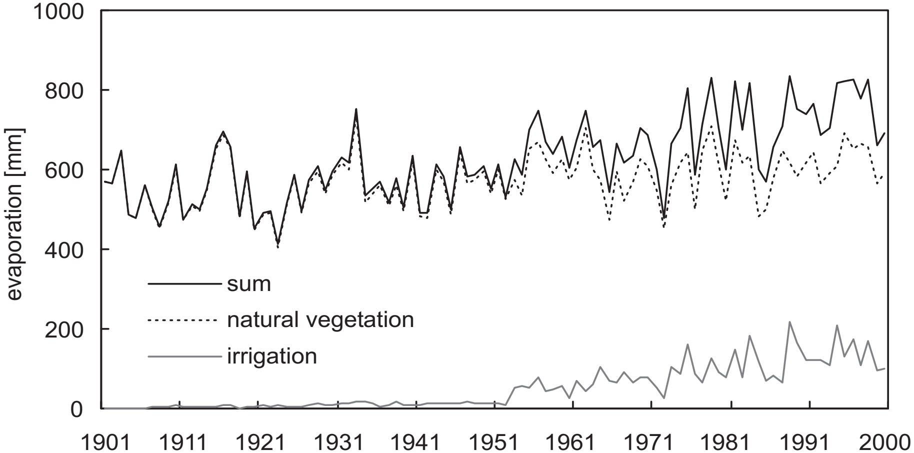

Fig. 8. Simulated evaporation by the natural vegetation, estimated additional water consumption for irrigation, and their sum.

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